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REPRINT
to
Unclassified Technical Reports
Distribution List of the
Geophysics Branch of the
Office of Naval Research

14 Contribution no. 1436; Ref[redacted] No. 64-20

(6) Selected Reprints of Atlantic Oceanography.

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(3) Some Biological Characteristics of the Marine Environment,

Bostwick H. Ketchum

The W.H.O.I. contribution to this work constitutes a technical report to the Office of Naval Research under Contract Nonr-2196(00), NR 083-004.

May 1964

APPROVED FOR DISTRIBUTION Bostwick H Ketchum
Bostwick H. Ketchum, Associate Director

Deep-Sea Research, 1964, Vol. 11, pp. 1 to 10. Pergamon Press Ltd Printed in Great Britain.

The sill depth of the Mid-Atlantic Ridge in the equatorial region*

W. G. METCALF†, B. C. HEEZEN‡ and M. C. STALCUP†

(*Received 5 December 1963*)

PREFACE

This report is a companion paper to that of Heezen, Bunce, Hersey and Tharp entitled "Chain and Romanche Fracture Zones," also appearing in this issue, which concerns the topography of the equatorial portion of the Mid-Atlantic ridge. The interpretation of the topographic data in that paper depended in some part upon the interpretation of the hydrographic station data which is discussed in the present paper, and *vice versa*. For this reason, it seems appropriate to publish both papers simultaneously in the same journal.

The lengths of the two reports dictated against presenting all the material in a single paper, but the authors wish especially to call attention to the inter-disciplinary dependence of the two portions of the study of the region involved.

Abstract—Recent evidence confirms earlier theories that the deepest communication between the Eastern and Western Basins of the Atlantic Ocean lies in the region of the Romanche Trench where the Mid-Atlantic Ridge crosses the Equator. New topographic data presented elsewhere suggests that the path of deepest communication is along either the Romanche or the Chain Fracture Zone. Hydrographic station data support this and indicate that the controlling depth is about 3750 m, which is from 250–1000 m shallower than earlier estimates. Various factors including scatter of the temperature and salinity characteristics in the ocean, internal waves and seasonal or longer period surges associated with the formation of Antarctic Bottom Water in the Western Basin combine to make the effective oceanographic barrier deeper than the true topographic barrier. Scatter due to inaccuracies in the sampling and measuring techniques have the further effect of making the oceanographic barrier appear deeper than it otherwise would.

Possible paths of flow of the deep water from Western to Eastern Basin are shown.

INTRODUCTION

A PROBLEM which has long interested oceanographers concerns the geographical location and the depth of the sill in the Mid-Atlantic Ridge separating the deep waters of the Eastern and Western Basins of the Atlantic Ocean. In the equatorial region within the Eastern Basin, there are traces of deep water whose relatively low temperature, salinity and oxygen definitely point to the Western Basin as their source. Using the geographical terminology given in SVERDRUP, JOHNSON and FLEMING (1942) we are concerned here primarily with the Brazil Basin in the west and the Cape Verde, Sierra Leone and Guinea Basins east of the Ridge. For convenience, the terms Western and Eastern Basins will generally be used in this report except where it is important to designate the individual basin.

*Woods Hole Oceanographic Institution Contribution No. 1436.
Lamont Geological Observatory Contribution No. 683.
†Woods Hole Oceanographic Institution, Woods Hole, Massachusetts.
‡Columbia University, Department of Geology and Lamont Geological Observatory, Palisades, New York.

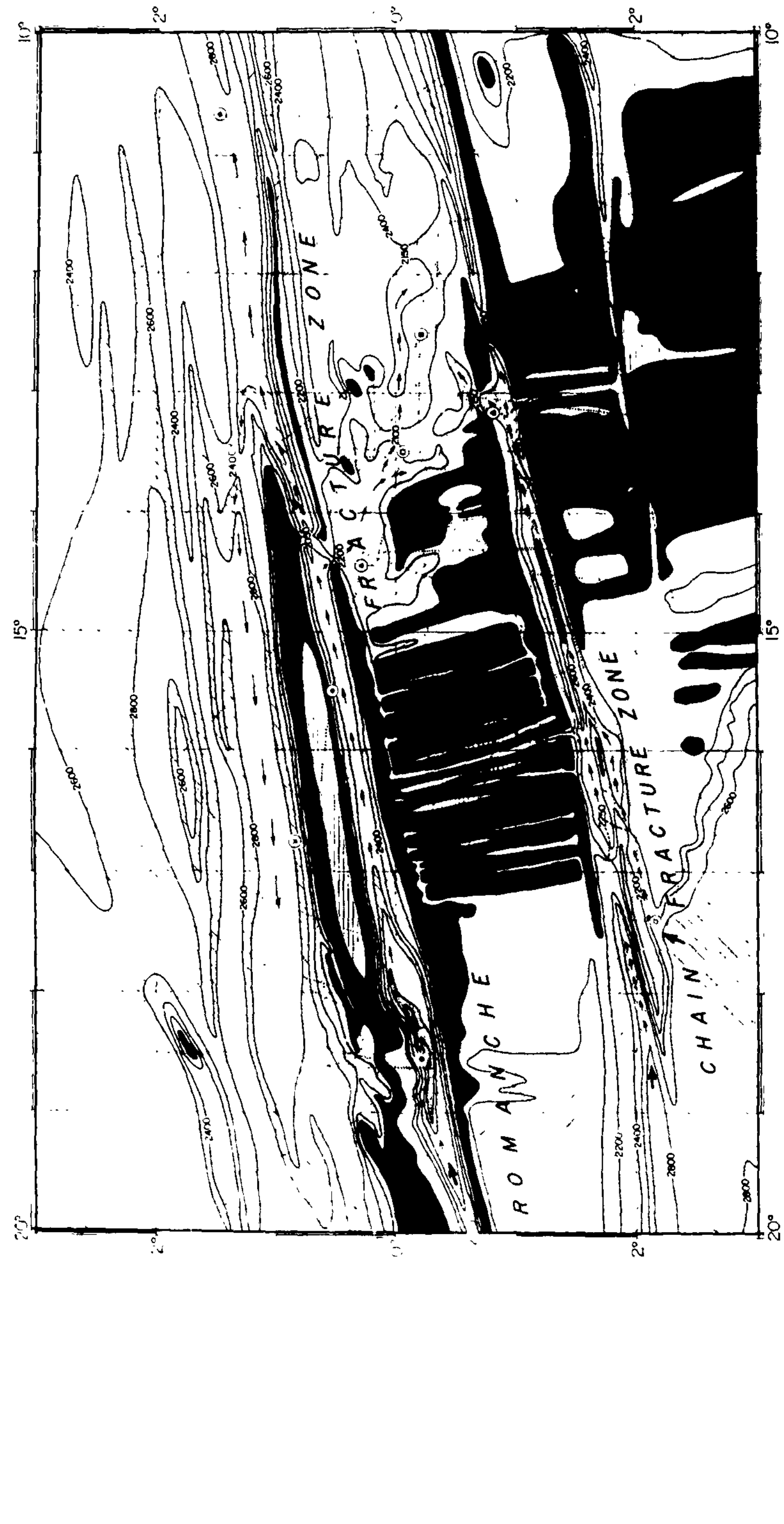

Fig. 1. Bathymetric chart of the Romanche and Chain Fracture Zones after HEEZEN, BUNCE, HERSEY and THARP (1964). Soundings are in uncorrected fathoms. Contours by Heezen. Arrows indicate possible flow of the deepest water based on topographic evidence combined with deep temperature observations. Positions are shown of the stations appearing in Figs. 3 and 4.

It is clear from the *Meteor* Atlas (WÜST and DEFANT, 1936) and the more recent surveys, especially those made in connection with the International Geophysical Year of 1957–1958 (FUGLISTER, 1960; WORTHINGTON and METCALF, 1961) that in the Eastern Basin, the deep water is cooler at the Equator than it is to either the north or the south. It has long been speculated that the region of the Romanche Trench, a remarkably deep feature on the Mid-Atlantic Ridge close to the Equator at about 18½°W long. might be where the deepest communication between the two basins takes place (DEFANT, 1927; BÖHNECKE, 1927; WÜST, 1936).

The bathymetric data previously available from the area have been so incomplete as to give only the vaguest hints as to the exact location and depth of the sill. Similarly, the temperature, salinity and oxygen data have provided only meager and ambiguous clues to the answer, being widely scattered in time, uneven in quality and frequently very uncertain as to geographical position.

GROLL (1912) in discussing his bathymetric chart gives frank warning that because of the sparsity of the data at his disposal the contours he drew were based on the broadest interpolations. His chart shows several 'breakthroughs' of the 4000-m isobath across the Ridge : one at the Romanche Trench and others at about 2°–3°S, 15°W.

BÖHNECKE (1927) believed that a single passage deeper than 4000 m exists in the Romanche Trench region and that no passage at this depth occurs to the south and east. WÜST (1936) accepted this latter view of the location of the sill on the basis of hydrographic data from the *Meteor* expedition and earlier cruises. He emphasizes the scarcity of the data and the uncertainty of his estimate which places the sill depth between 4500 and 4800 m.

The recent surveys by *Vema* (Cruise 12, 1957) and *Chain* (Cruise 17, 1961) which have been described at length by HEEZEN, BUNCE, HERSEY and THARP (1964) while not giving the final answers as to the depth and location of the sill of the Mid-Atlantic Ridge, have gone a long way toward narrowing the field. As HEEZEN, BUNCE, HERSEY and THARP comment, " because of strong surface currents and frequent cloudiness, errors in navigation can be relatively large." Nevertheless, the bathymetric chart constructed by those writers and reproduced here as Fig. 1 is by far the most detailed and accurate one produced up to the present time. The soundings shown in the figure are in fathoms as read from the echo sounder and are corrected only for the depth of the echo sounding transducer, not for the calculated speed of sound in sea water. The hydrographic data are given in corrected meters, conversions from uncorrected fathoms being made according to MATTHEWS' tables (1939).

In spite of the extensiveness of the bathymetric survey, there are, as is usually the case in this sort of study, several uncertainties in the interpretation of the material. In the construction of the chart, use was made of the hydrographic station data. The temperature and salinity characteristics of the deep water, differing as they do in the two basins, give important clues as to the controlling depths of the topographic features lying between the individual or groups of stations in the area studied. These clues are considered in the section which follows.

DISCUSSION

Hydrographic stations from the *Chain* cruise in 1961 and from the Equatorial Section of an I.G.Y. cruise by *Crawford* in 1958 (METCALF, 1960) were grouped

according to which of the basins they represented. A total of 24 Eastern Basin stations and 11 Western Basin stations were used. Many more stations were available from both sides of the Mid-Atlantic Ridge in this general region, but it was felt that it was desirable to have all the data as closely comparable as possible with respect to equipment and technique. The two cruises were alike in these respects.

In Fig. 2a, the potential temperatures below 3400 m are plotted versus depth, and envelopes encompassing Eastern and Western Basin observations were drawn. As is shown in the figure, the two envelopes diverge at a depth of about 3750 m. (For comparison with Fig. 1, this equals 2000 fm, uncorrected for the speed of sound). Below this, the Western Basin potential temperatures are markedly cooler at all depths than those in the Eastern Basin.

Similarly, as shown in Fig. 2b, the salinity observations were plotted versus depth. Again there is a divergence, the deep salinity in the Western Basin being lower than those in the Eastern Basin. The divergence of the two envelopes in this case is at about 3925 m (2100 fm, uncorrected). The difference in the depths of the divergences of the two characteristics is discussed below.

Arithmetical average temperatures were computed for both groups of stations from interpolated values at 50-m intervals, and standard deviations from the mean were calculated. It was found that a curve constructed from the values of *plus twice the standard deviation* from the Western Basin averages diverged from a curve representing *minus twice the standard deviation* from the Eastern Basin averages at approximately 3750 m. In other words, below that depth the separation between the temperature characteristics of the two basins can be considered to be of practical significance.

The question arises as to exactly what is the relationship between the level of divergence of these curves and the depth of the true physical barrier between the Eastern and Western Basins. It seems clear that the physical barrier is no deeper than 3750 m, and it may be shallower.

Scatter in the accuracy of the temperature data affecting both the apparent temperature and the computed depths (depths were calculated from paired protected and unprotected deep-sea reversing thermometers) would increase the spread of the envelope of observations in each basin which in turn would have the effect of making the apparent separation between the two basins deeper than it should be. Actual scatter in the true temperature/depth relationship in the ocean would have the same effect.

It is customary to consider temperatures from properly functioning reversing thermometers to be accurate to about $\pm$ 0·02°C. This takes into account various factors such as the precision of the thermometers, the accuracy of the calibration, the care and skill with which they are read and the accuracy of the temperature correction techniques, among other things (WHITNEY, 1957). For the temperature/depth averages in the Eastern Basin, twice the standard deviation from the mean was frequently as low as 0·02 suggesting that if we can expect that much scatter in accuracy of the thermometric readings, then there is practically no scatter whatsoever in the deep ocean temperatures in this region. It would seem more likely that a small amount of scatter is present in the ocean, and that our thermometers and techniques are better than we have been giving them credit for being.

Another factor involved here in the relationship between the true physical sill depth and the apparent sill depth as derived from temperature and salinity measure-

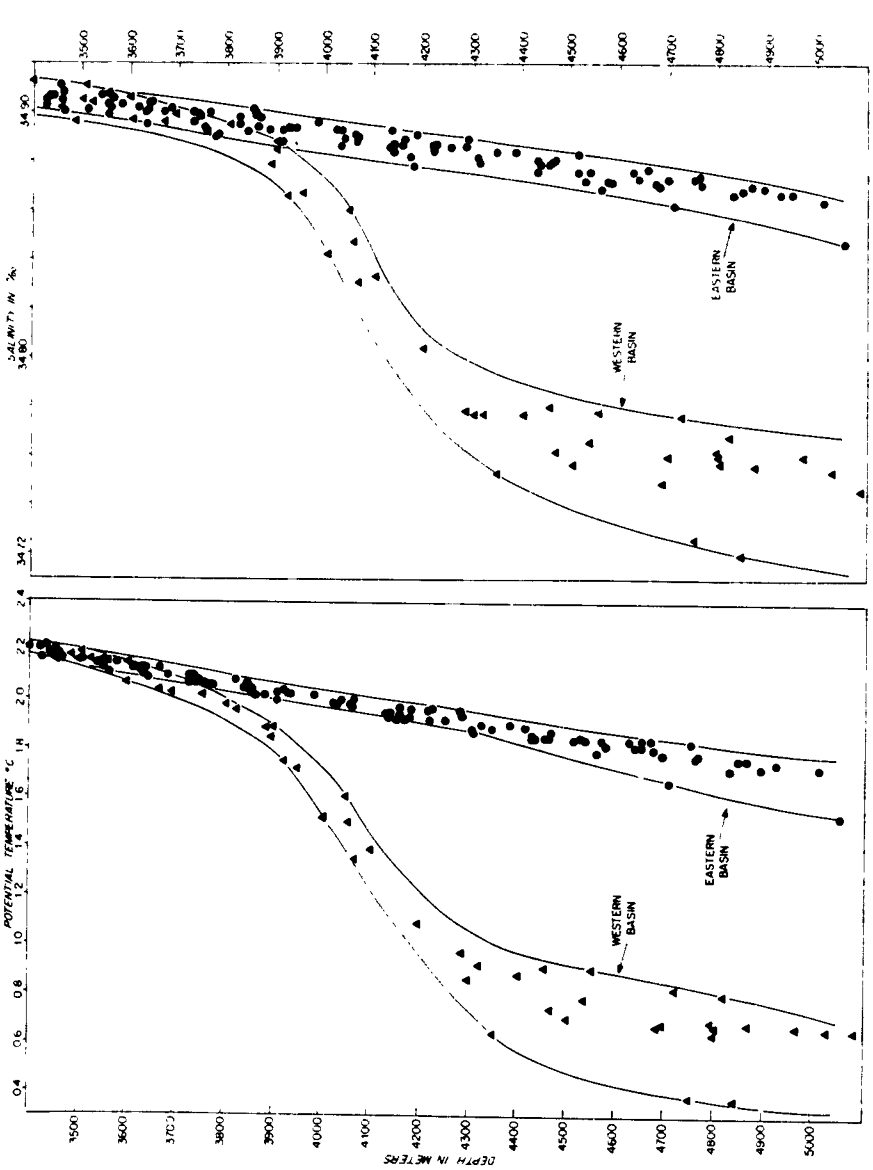

Fig. 2. Potential temperature/depth (a) and salinity/depth (b) relationship in the deep water of the Western and Eastern Basins of the equatorial Atlantic based on *Crawford* and *Chain* hydrographic stations. The separation of the envelopes suggests the effective depth of the sill of the Mid-Atlantic Ridge. The discrepancy between the depths of separation shown in the two portions of the figure is discussed in the text.

ments has to do with vertical water movements in the sill area. Presumably the cold deep water in the Western Basin is renewed periodically in the Antarctic. When any sizable volume of this cold dense water is formed and subsides to the bottom, the water layers under which it comes to lie will be raised somewhat. Following a suitable lag period for the new deep water to spread across the basin, the effect would be felt in the vicinity of the sill, and water which had been lying below sill level would be lifted to the point where it would be able to flow over the sill. This would have the effect of making the sill depth deduced from the separation of the temperature or salinity characteristics in the two basins appear deeper than it is in reality unless, of course, the measurements happened to have been made at the time of the greatest rise in level of the Western Basin deep layers. In this connection it should be pointed out that the *Crawford* stations occupied in November 1958 and the *Chain* stations occupied in March and April 1961 show no regular differences in their temperature/depth or salinity/depth relationships in this region.

By the same line of reasoning, any internal waves which may be present in the region would have the effect of making the oceanographic sill deeper than the true bathymetric sill by the amount of the wave amplitude. At present there is no information as to how much this factor may amount to.

As was touched on above, a puzzling aspect of the problem arose when the method used in dealing with the temperature data was used in an attempt to deduce the sill depth independently using the salinity data. As is shown in Fig. 2, the depth at which the envelopes diverge is considerably greater in the case of the salinity data than in that of the temperature data : 3925 m compared with 3750 m. Probably the explanation lies in the fact that salinity measurements at the present state of the art simply are less accurate than are the temperature measurements. This would cause a broadening of the salinity envelopes as compared with the temperature envelopes thus causing them to diverge at a deeper level. Therefore we feel that the evidence of the sill depth derived from the temperature data is more realistic than that derived from the salinity data.

This matter of salinity measurements being less accurate at the present state of the art is based on the following reasoning. At present, we are measuring temperature reasonably accurately to 0·005°C and salinity to 0·005‰. In the deep water with which we are concerned (around 2°) a difference of temperature of 0·005° amounts to a sigma-t change of 0·0004, assuming a salinity of about 35‰. On the other hand, a difference in salinity of 0·005‰ at 2° amounts to a change in sigma-t of 0·004; an order of magnitude greater.

In other words, as far as the effect on the physics of the ocean is concerned, our salinity measurements are much rougher than our temperature measurements. It is felt that this roughness, which really amounts to scatter in accuracy, is effective in broadening the salinity envelope as compared with the temperature envelope in the present case, thus forcing the separation deeper.

Similarly, the observations of the dissolved oxygen, which show a much greater scatter than do those of temperature and salinity, lead to deductions of the sill level at an even greater depth.

When the stations were being grouped as to Western or Eastern Basin type, it was apparent that a number of distinct sub-types could be identified. These would resemble one or the other major type down to some particular depth below the postu-

lated sill before diverging from the rest of the group. This appeared to be connected to their geographical positions relative to the path of deepest communication between the two basins.

Figure 3 is an attempt to illustrate schematically a series of stations and the effect of topographic barriers separating the deeper portions of their water columns. The Western Basin is shown at the left and the Eastern basin at the right with various areas of the Mid-Atlantic Ridge between. The spacing is not intended to represent relative distances, and the sequence of the stations is not intended to imply that the stations are located along a direct flow from one basin to the other. The actual locations of the stations are shown in Fig. 1.

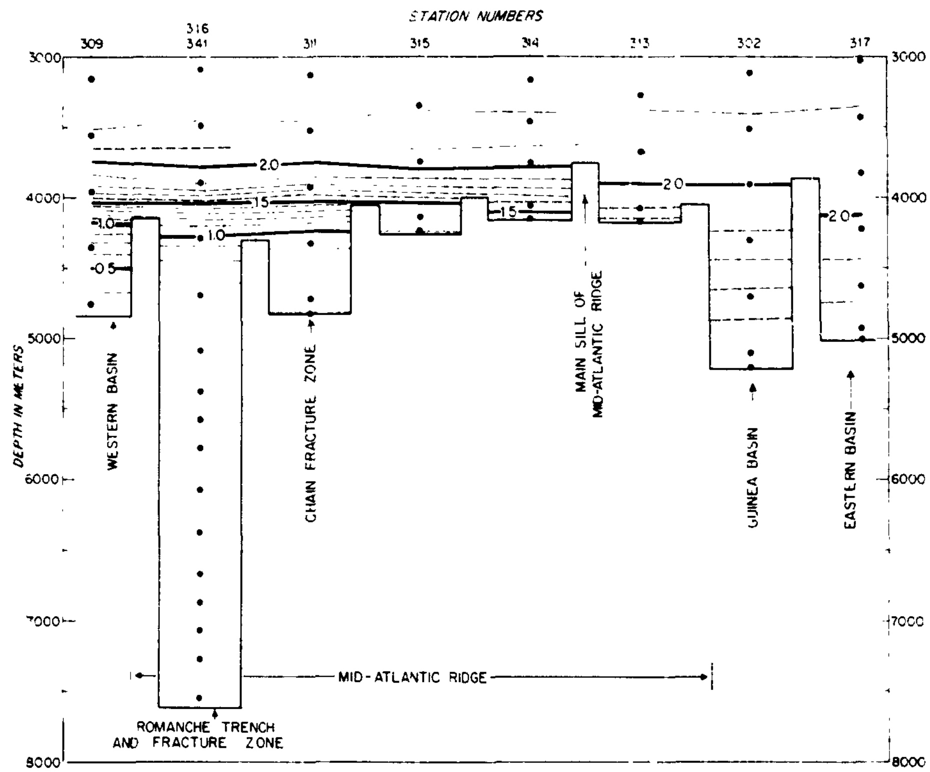

Fig. 3. Schematic profile of potential temperature across the Mid-Atlantic Ridge. The spacing and order of the stations are not intended to represent relative distances nor sequence of stations in a direct flow from one basin to the other. See Fig. 1 for station positions.

The stations in the Romanche and Chain Fracture Zones resemble the Western Basin station down to about 4070 m (2170 fm, uncorrected) which may represent the depth of the effective oceanographic barrier between them and the Western Basin. A barrier at 4300 m (2290 fm, uncorrected) appears to separate the two fracture zones. The stations on the Mid-Atlantic Ridge appear to be in minor pot holes or culs-de-sac which separate their deepest water from water at the same depth in the nearby fracture zones and from each other. Stations 314 and 315 appear to be west of the main sill,

and Station 313 is east of it but separated from the deeper Guinea Basin and other parts of the main Eastern Basin by a barrier.

Station 302 in the Guinea Basin appears to be close to the eastern outlet of the deepest flow across the sill and shows the coldest deep water we have found lying definitely east of the Ridge. This station is cooler than others also found in the Guinea Basin which may indicate that it lies in a depression separated from the main part of

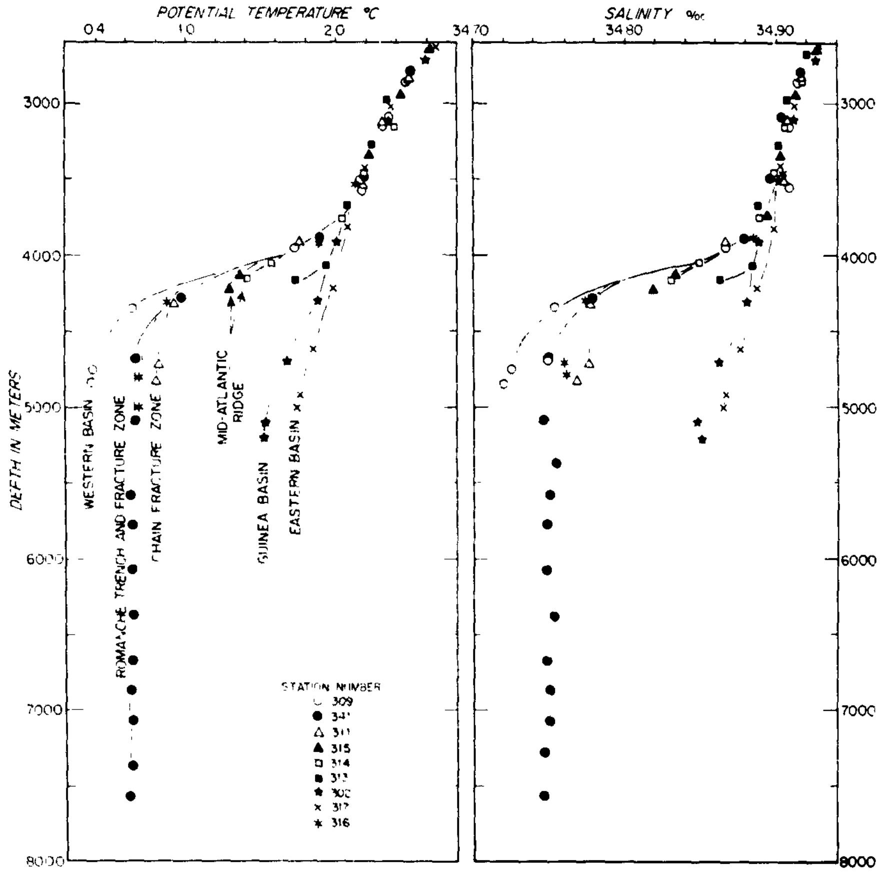

Fig. 4. Comparison of the potential temperature/depth (a) and salinity/depth (b) relationship of the stations shown in Fig. 3. See Fig. 1 for station positions. The points of division between the various curves are believed to be related to the topography separating the corresponding stations.

that basin and from the Sierra Leone Basin by a ridge. Or it may mean only that as the water enters the Guinea Basin it is very quickly modified by mixing with the surrounding water.

Figure 4 (a and b) compares in another form the potential temperature (a) and salinity (b) of the water columns of this same series of stations. Again it should be emphasized that this is purely a schematic representation comparing selected stations and is not intended to suggest that any of these stations are in a direct line of flow between the two main basins.

From this material, we are now in a position to suggest possible lines of flow of deep water through this complex area. Referring again to Fig. 1, we wish to call

attention to the arrows which give our interpretation of the deep flow as deduced from the topographic and physical oceanographic data at hand.

Station 309 is used as the ' type station ' for the Western Basin. As was seen in Figs. 3 and 4, stations in the Romanche Trench and Fracture Zone, Nos. 341 and 316 respectively, and in the Chain Fracture Zone, No. 311, differ from the Western Basin type only slightly and only in the very deep part of their columns.

Stations 315, 314, and 313 show progressively greater modification from the Western Basin pattern. The fact that No. 313 shows the greatest modification although it is geographically closest to the Chain Fracture Zone suggests that any flow from that Zone is minor. These three stations appear to be more or less in a line of flow coming from the Romanche Fracture Zone. However, observations in the deep basin further east indicate from their high temperature at the bottom that little if any of this flow gets all the way across the Ridge in this region.

Station 302, which is the coldest of the stations in the deep water east of the Ridge, is much further away, geographically, from the presumed source of flow through the Romanche Trench than is the much warmer station 317 directly north of the Trench. The structure of the Romanche Fracture Zone, however, makes it clear that No. 302 is much closer to the source along the round-about route which the water is forced to take in reaching the region of No. 317.

SUMMARY

The various factors causing vertical water movements in the sill area, producing scatter in the temperature/depth and salinity/depth relationship in the ocean, and affecting the accuracy of the temperature and salinity observations all tend to make the sill depth appear greater than it really is. Therefore a system, assuming it is scientifically acceptable in other respects, which indicates the shallowest sill is considered the most promising one. The measurement of temperature in the ocean is undoubtedly more reliable than the measurement of the other commonly observed characteristics, and therefore it is considered significant that the system involving the temperature separation between the two basins should be the system suggesting the shallowest sill depth. In other words, it is felt that the figure of 3750 m for the maximum controlling depth of the sill is the most realistic one to date.

However, it must be kept in mind that the true depth of the sill may be appreciably shallower than the controlling depth due to the various causes of water movements discussed. How much shallower than 3750 m the physical sill may be is something about which we have little information at this time.

Physical oceanographic data combined with the topographic data have been used to deduce a possible route of supply of Western Basin water across the Mid-Atlantic Ridge. It appears that the major flow passes through the Romanche Trench and Romanche Fracture Zone and that the Chain Fracture Zone carries little if any water through to the eastern side of the Ridge. Various hydrographic stations in the area suggest that they are affected to some extent by the Western Basin cold deep water but are not on the direct path.

Acknowledgement This work was supported by the Office of Naval Research under Contract Nonr-2196 with the Woods Hole Oceanographic Institution and Contract Nonr-266 (48) with Columbia University. The assistance of G. LEONARD JOHNSON is gratefully acknowledged.

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(*From the* Quarterly Journal of the Royal Meteorological Society, Vol. 90, No. 384, April 1964)

Sea smoke and steam fog

By P. M. SAUNDERS
Woods Hole Oceanographic Institution, Massachusetts

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Sea smoke and steam fog

By P. M. SAUNDERS
Woods Hole Oceanographic Institution, Massachusetts

(Manuscript received 9 July 1963; in revised form 29 January 1964)

SUMMARY

The characteristics of fogs resulting from the advection of cold air over warm water (steam fog and sea smoke) are investigated. They are found to occur with air temperatures between 5 and 40°C lower than the water temperature, in winds from calm to gale; they have liquid water contents in the range 0·01-0·5 g m^{-3}, extend in height from 1 to 1,500 m and commonly exhibit either a columnar or banded structure. A study of their occurrence in Atlantic waters reveals a marked concentration in the western regions in the winter months because of the proximity of warm ocean and cold continent.

Through the use of equations for turbulent transfer it is shown that the occurrence of steaming is related to the well-known fact that two masses of unsaturated air at different temperatures when mixed together *can* yield a supersaturated or foggy mixture. In deriving the connexion the equality of transfer coefficients for heat and water vapour is assumed.

The circumstances in which steaming occurs are defined. The difference in the temperatures of the air and water must exceed a threshold which is dependent on the humidity of the air and the temperature and salinity of the water : its value, in the range 5-15°C, is a minimum when the air is moist and the water cold and fresh. The liquid water content and vertical extent of steaming increase as the thermal contrast increases. Close agreement is found between observations of the onset of steaming and the computed threshold values.

1. INTRODUCTION

The exchange of heat and vapour from a wet surface into the air overlying it is sometimes accompanied by 'steaming.' Wet, sun-warmed soil may steam; so may a body of water if the air above it is sufficiently cold. In the latter case the phenomenon is commonly termed steam fog (over fresh water) or sea smoke (over saline water)*. The major concern here is to define the circumstances in which steaming occurs : this is accomplished in Section 3 of the paper and shown to be correctly argued in Section 4. The theoretical discussion and examination of data is preceded by a summary of the distribution and physical characteristics of steam fogs since no current text is found to treat the topic in anything approaching a comprehensive manner.

2. DISTRIBUTION AND PHYSICAL PROPERTIES OF SEA SMOKE AND STEAM FOG

The steaming of natural waters takes place on all meteorological scales; micro, meso and synoptic. It occurs on a micro-scale when air cooled by nocturnal radiation drains from high ground onto a pond (Horton 1933; Woodcock and Stommel 1947). It occurs on a meso-scale when cold air cascades off the dark chill land mass of northern Norway onto the ice-free areas of the Fjord waters (Spinnangr 1949). It occurs on a synoptic scale when, in the wake of a depression, continental arctic air sweeps off the north-eastern coast of the U.S.A. out over the adjacent Atlantic (Brooks 1934).

(a) *Distribution of sea smoke (N. Hemisphere)*

Some features of the distribution of steaming on the synoptic scale were found by plotting about 60 reports of steaming in the Pacific and Atlantic Oceans. The following generalizations may be made : (*i*) Sea smoke commonly occurs outside polar latitudes and has been reported as far south as the tropics (Bannister 1948; Starbuck 1953) : thus the term arctic sea smoke seems particularly inappropriate; (*ii*) Over 90 per cent of the

* Other terms for the phenomenon are listed in the Appendix.

observations of sea smoke were made in the winter months, December to March; (*iii*) Because of the presence of warm water close to the western oceanic boundaries and because of the general eastward motion of cold continental air, the western parts of the ocean experience more frequent and more intense steaming than the corresponding eastern parts; (*iv*) Steaming is generally confined to coastal waters in low latitudes but with increasing latitude is reported at locations increasingly remote from the coast : Hay (1953) describes sea smoke at latitude 60°N after the air had an overwater trajectory of nearly 1,000 miles.

(*b*) *Factors influencing their occurrence*

Because it is well known that only when overlying air is *sufficiently* cold can a water surface steam, many investigators have looked for a threshold value for the difference in temperature between air and water. Thus Jacobs (1954) asserts that steaming never occurs in the Gulf of St. Lawrence when the air is less than 9°C colder than the water. Yet Horton (1933) observed steaming of a pond with a difference of only 6°C, and the author (see Table 1) has observed differences near 14°C in the absence of sea smoke. As demonstrated later, the relative humidity of the cold air is an important factor in this threshold value.

Observations of steaming have been made in winds ranging from near calm (Bryson 1955) to nearly 30 m sec^{-1} (Rodewald 1937; Spinnangr 1949), but according to Church (1945) its speed has little effect on whether a water surface steams or not.

(*c*) *Their vertical extent*

Steam fog and sea smoke are widely regarded as shallow phenomena : they are not. In deep cold air the height of sea smoke can exceed 1,500 m (Jacobs 1954; Berry, Bollay and Beers 1945; Cunningham (private communication)), and ship reports of steaming in excess of 100 m are not uncommon. The depth of steaming may on occasions be limited by the vertical depth or stratification of the cold air.

(*d*) *Their form*

The characteristic form of steam fog and sea smoke varies with the wind. In near calm Bryson (1955) observed an array of quasi-steady convergent columns with a spacing of a few meters and height of 2 m. Horton (1933) describes similar columns which were rotating; these had a diameter of ½ m and height 5 m and drifted unsteadily across a river. On a larger scale, Brooks (1934) Woodcock (private communication) and others have described unsteady rotating columns of fog with vertical dimensions of 100 m or more, the scene likened to Dante's *Inferno*; Woodcock reported a wind of 5 m sec^{-1}. Church (1945) has used the terms sheet and blanket, indicating a well-marked top to the fog layer. The author's observations indicate that in moderate winds sea smoke commonly has a banded structure (see Fig. 1, Plate V) with the bands approximately along the wind, and it is a surprise to find only one other mention of this form in the literature (*Marine Observer* 1931, **8**, p. 60). In deep cold air steaming may be accompanied by cumulonimbus (*Marine Observer* 1957, **27**, p. 187; 1960, **30**, p. 12), cumulus or low stratocumulus.

(*e*) *Visibility and water content*

In sea smoke visibility as low as 30 m (*Mar. Obs.* 1959, **29**, p. 11), 50 m (Brooks 1934) and 100 m (Rubin 1958) has been reported : despite the widespread use of radar such obscuration represents a navigation hazard. On the other hand, steaming may be so slight and shallow that the visibility is not appreciably affected : in this case it is common to observe pronounced refractive shimmering and cusping of the horizon.

From the observations of visibility we can estimate the water content in steam fog as up to several tenths of a gram per cubic metre (Houghton and Radford 1938). The estimate is confirmed by direct observation of water content in winter fogs over the river Angara (in an industrial area) by Bashkirova and Krasikov (1958); these authors made a rough classification of fogs into tenuous, moderate, and dense, and found water contents of 0·03-0·04 g m^{-3}, 0·05-0·11 g m^{-3} and 0·08-0·37 g m^{-3} respectively. The water content was found to increase with increasing air-water temperature difference. In the cleaner air of the Arctic, in Kola Bay near Murmansk, water contents of 0·02-0·04 g m^{-3} and 0·04-0·14 g m^{-3} were reported in moderate and dense fogs.

(*f*) *Microphysical structure*

(1) Phase state

At air temperatures below 0°C the condensed water in sea smoke and steam fogs is commonly supercooled : riming of exposed surfaces in dense cold steam fog may thus be considerable (see Lee 1958; Mitchell 1958; and foreground, Fig. 1, Plate V). The Russian investigators of the steaming Angara River found that at temperatures above − 9°C to − 10°C the fogs consist of supercooled drops alone, whilst at temperatures below about − 20°C the condensate was predominantly ice. At intermediate temperatures fogs were mixed, with many spherical frozen drops. On occasions of steaming of Kola Bay, fogs remained supercooled to much lower temperatures, − 18°C to − 22°C, before crystals and irregular solid particles formed. Bashkirova and Krasikov interpret the differences as due to the industrial contamination of the Angara fogs.

(2) Drop sizes

One important feature of both the Angara River and Kola Bay fogs was that a decrease in air temperature was accompanied by a *decrease* in the size of the most frequent drop and a *decrease* in the width of the spectrum but an *increase* in water content; a similar behaviour was exhibited by the solid phase. Thus a reduction of air temperature resulted in the activation of not only greater numbers of freezing nuclei but also greater numbers of condensation nuclei. The writer believes that the latter result is a reflection of the increase in the rate at which air is brought to the condition of saturation with increasing air-water temperature difference.

3. The joint transfer of heat and water vapour

(*a*) *A necessary condition for the onset of steaming*

The simplest circumstances of steaming arise when ***deep homogeneous*** cold air overruns *fresh* warm water of *uniform* surface temperature. The potential temperature of air crossing the shore is denoted as θ_0, its mixing ratio as r_0. There is no restriction of the shape of the shoreline nor the variation of wind with height or time except in order to preserve unambiguity of the phrase ' downwind of the shore.' The equation for the *mean* value of a transferable property x is then written with the usual notation :

$$\frac{dx}{dt} = \frac{\partial}{\partial x}\left(K_x \frac{\partial x}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y \frac{\partial x}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z \frac{\partial x}{\partial z}\right) \quad . \quad . \quad (1)$$

where x is θ and r in turn, and K_x, K_y, K_z are the three components of the turbulent transfer coefficient. At the water surface ($z = 0$) it is supposed that the air takes up the

Figure 1. Sea smoke in Great Harbour, Woods Hole, Mass., U.S.A. 0915 EST, 31 December 1962. Water temperature $-0{\cdot}2$°C; air temperature (10 m) -15°C, humidity 60 per cent. Wind NW 10 m sec^{-1}. Height of steaming approximately 5 m.

temperature T_F of the surface and the saturation mixing ratio for this temperature $r_v(T_F)$. Hence boundary conditions are

$$\theta = \theta_0, \qquad r = r_0 \qquad \text{at shore and} \qquad z > 0$$

$$\theta = \theta_F = T_F, \; r = r_v(T_F) \text{ downwind of shore and } z = 0 \qquad (2)$$

If the transfer coefficients for heat and water vapour are everywhere equal then it follows from Eqs. (1) and (2) that θ and r are everywhere linearly related; that is

$$r - r_0 = \beta(\theta - \theta_0) \qquad (3)$$

where

$$\beta = \frac{r_v(T_F) - r_0}{\theta_F - \theta_0} \text{ is a constant} \qquad (4)$$

This simple and powerful result holds however complex is the dependence of wind and transfer coefficients on space and time; it fails, however, if the transfer coefficients for heat and water vapour are unequal. Although differences have been reported (see Priestley 1959) they are believed to vanish close to the surface where the Richardson number is small. Since the major fraction of the temperature-drop in cold air over warm water occurs (in the mean) below a height of about 10 cm, the assumption of equality should be good. The roles of molecular conduction and diffusion, whose coefficients are also unequal, are supposed confined to an extremely shallow layer immediately adjacent to the water surface.

Eqs. (3) and (4) can be identified with the mixing law for two moist air masses with characteristics r_0, θ_0 and $r_v(T_F)$, θ_F – the latter to be interpreted as resulting from intimate contact of air with the water surface. Then, as is well known, the result of mixing the two air masses is found in the r, θ plane on the straight line joining r_0, θ_0 and $r_v(T_F)$, θ_F – as is stated in Eqs. (3) and (4).

For fresh water the point $r_v(T_F)$, θ_F also lies on the curve of saturation mixing ratio versus temperature and reflection shows that the tangent to the curve at this point divides the r, θ plane in a significant manner. If the point r_0, θ_0 lies below the tangent, steaming of the surface cannot take place, if r_0, θ_0 lies above the tangent, steaming may take place. For in the latter case, r_0, θ_0 in the hatched area of Fig. 2, some of the air modified by mixing acquires a mixing ratio which exceeds the saturation value for its temperature; this we suppose to be a necessary condition for steaming. Thus for steaming

$$\beta < \left(\frac{dr_v}{dT}\right)_{T_F}. \qquad (5)$$

and for *just no steaming*

$$\beta = \left(\frac{dr_v}{dT}\right)_{T_F}. \qquad (6)$$

Given the initial condition of the cold air, Eqs. (4) and (6) permits the determination of the value of the water temperature for just no steaming. (From the graphical interpretation it is clear that Eq. (6) only possesses a solution if $d^2 r_v/dT^2 > 0$. Hutton (1788) in considering why the breath of animals is sometimes rendered visible correctly inferred that 'the solution of water in air increases with heat (temperature) in an increasing rate.' He then proceeded to a theory of clouds and rain based on mixing. Because of the observed complex dependence of r_v on T the determinations must be made numerically; results are shown in Fig. 3. The values computed are threshold values in the sense that if the air is colder or has higher relative humidity than shown, steaming can occur. From Fig. 3 it is noted that the air-water temperature difference for just no steaming increases with *increasing* water temperature and has a minimum value when the air is moist and the water cold.

The presence of dissolved inorganic salts in the ocean results in a lowering of the equilibrium vapour pressure below that for fresh water at the same temperature. If $1 - f$ is the fractional depression then $r_s = f \cdot r_v(T)$ where r_s and $r_v(T)$ are the equilibrium values for contaminated and fresh water respectively. In the problem where deep homogeneous air overuns *saline* water of uniform temperature $T_s = \theta_s$, r and θ are again linearly related as in Eq. (3) but the slope β_s is now

$$\beta_s = \frac{f \cdot r_v(T_s) - r_0}{\theta_s - \theta_0} \qquad (7)$$

and the condition for just no steaming is

$$\beta_s = \lambda \left(\frac{dr_v}{dT}\right)_{T_s} \text{with } \lambda < 1 \qquad (8)$$

Examination of Fig. 2 should make clear the need of λ, for tangency is now required of the line joining r_0, θ_0 to r_s, θ_s with the curve of saturation mixing ratio versus temperature, where r_s no longer lies on the saturation curve. Evidently for a given condition of cold air the surface temperature of saline water necessary to promote steaming must be higher than that of fresh water and the difference is denoted as $\Delta\theta$. Using the fact that the saturation mixing ratio is accurately proportional to $\exp(\alpha T)$ over a limited range of temperatures, it may be shown that

$$f = \{1 + \alpha\Delta\theta\} \exp - \alpha\Delta\theta \qquad (9)$$

and

$$\lambda = \exp - \alpha\Delta\theta \qquad (10)$$

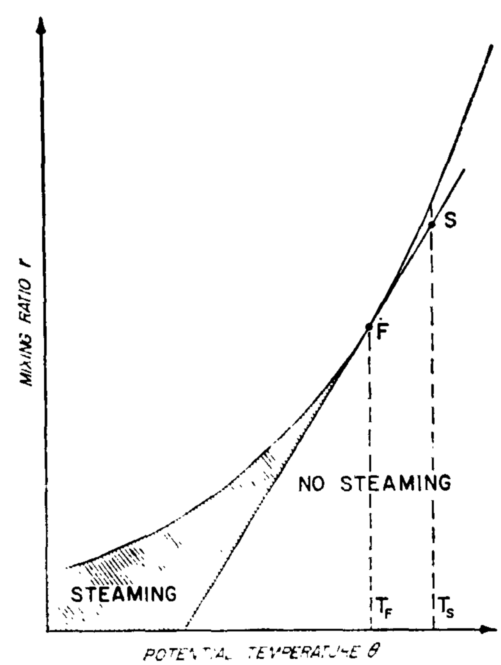

Figure 2. A criterion for the steaming of a fresh-water surface of temperature T_F and a saline-water surface of temperature T_S. SF is tangent to the saturation curve at the temperature T_F.

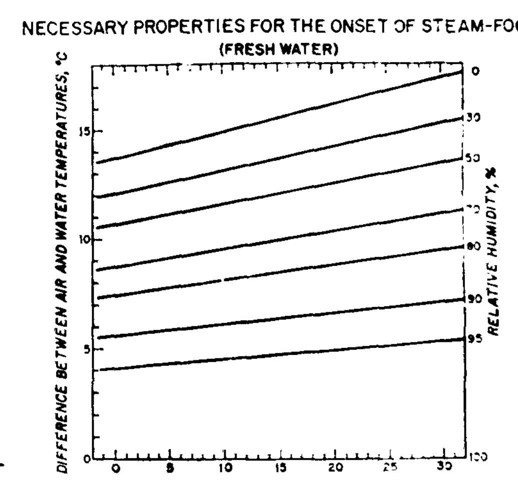

Figure 3. Necessary properties of air for the steaming of fresh water. The relative humidity is measured near the surface in the cold air.

Values of λ and $\Delta\theta$ as a function of f are shown in Fig. 4 where α has been allotted the values :

water temperature,	°C	0	15	30
α,	$°C^{-1}$	$7{\cdot}3 \times 10^{-2}$	$6{\cdot}5 \times 10^{-2}$	$5{\cdot}7 \times 10^{-2}$

The criterion for just no steaming is seen to be surprisingly sensitive to small amounts of impurities. For a depression of the equilibrium vapour pressure of only 0·01 per cent (salinity 0·2 per mille), a value reached in fresh-water lakes and rivers, $\Delta\theta$ is 0·2°C; but for a depression of 1·88 per cent (salinity 35 per mille), a value characteristic of the worlds ocean surfaces, $\Delta\theta$ is between 2·5 and 4°C ! This difference is so large that a diagram has been prepared showing the circumstances in which a saline surface is expected to steam (Fig. 5).

(b) *The liquid water content in sea smoke and steam fog*

When the characteristics of the cold air lie in the steaming region (Fig. 2), θ and r in Eq. (1) and subsequently are more logically interpreted as the wet-bulb potential temperature and the mixing ratio in both the liquid and vapour phase respectively (Rodhe 1962). However, in computing the water content here, a simpler procedure outlined by Brunt (1935) was followed.

Sample results are shown in Figs. 6 (a) and 6 (b) in conditions judged to correspond to weak and intense steaming of a saline surface. As is anticipated from the graphical interpretation of Eqs. (3) and (4) given earlier, increasing the thermal contrast between the air and water (for given r_0) increases both the water content of the fog and the *range* of air temperatures which sustain saturation. Even in extreme conditions, temperature contrast 40°C, the fog liquid water reaches a value of only about 1 g m^{-3}.

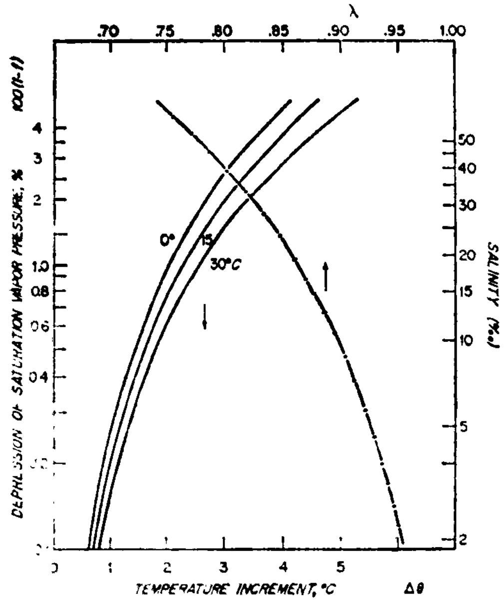

Figure 4. Difference between the surface temperature of contaminated water and of fresh water necessary to promote steaming for given cold air characteristics as a function of depression of saturated vapour pressure (solid lines); also λ of Eq. (5) as a function of the same depression (chain line).

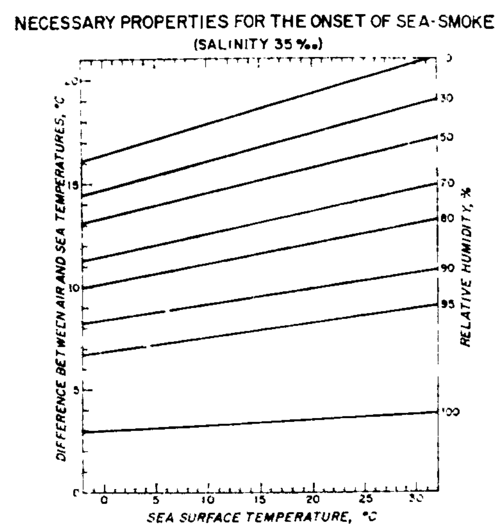

Figure 5. Necessary properties of air for the steaming of saline water (salinity 35‰). The relative humidity is measured near the surface in the cold air.

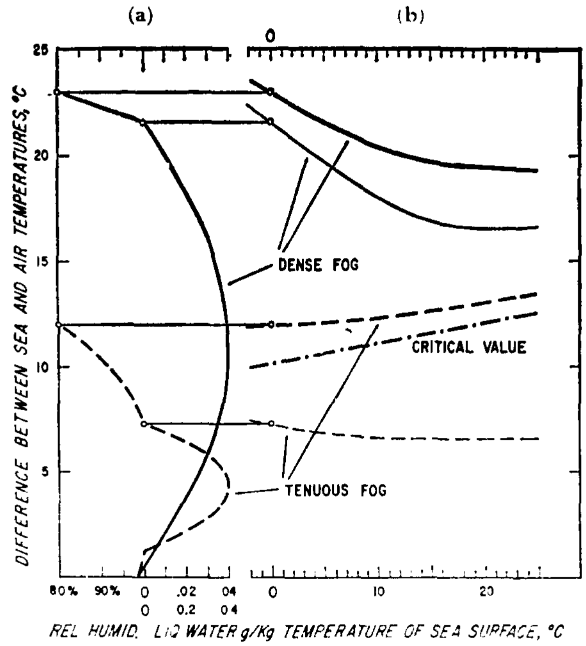

Figure 6. (a) Liquid and vapour in tenuous and dense fog; saline-water temperature 0°C and air relative humidity 80 per cent; (b) Variation of fog properties with water temperature.

4. A comparison between theory and data: conclusions

The following procedure has been adopted to test the simple ideas advanced in Section 3. Given an observation of steaming, the air temperature necessary for the onset of steaming is determined from the measured water temperature and measured relative humidity (Figs. 3 and 5); the difference between the measured and threshold air temperature is thus obtained. This difference is plotted against water temperature in Fig. 7 along with information about the vertical extent of steaming: occasions of steaming are distinguished from occasions of no steaming by the use of closed and open symbols respectively. The observations, which are drawn from reports by Church (1945), Hay (1953), *Marine Observers Log*, Rodewald (1937, 1959), Starov (1955) and Woodcock (private communication), show an approximate division between steaming and no steaming for air temperatures close to, but somewhat lower than, the theoretical values.

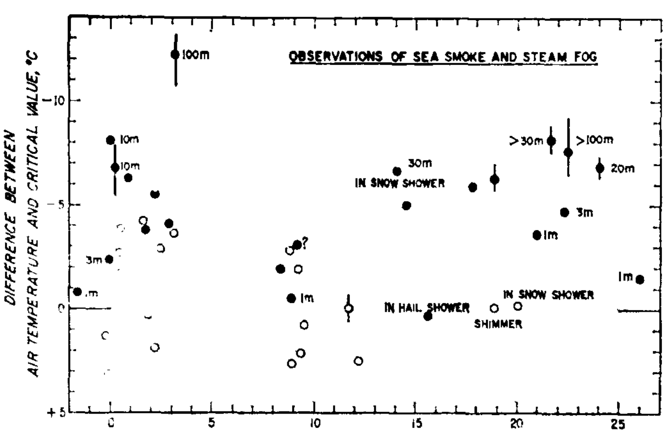

Figure 7. Reports of sea smoke and steam fog. For observers, see text, Section 4.

Careful investigation of the transition from steaming to no steaming has been made by the author on two occasions in the saline waters of Great Harbour, Woods Hole, Massachusetts. Measurements of dry- and (frozen) wet-bulb temperatures were made on the upwind shore of a peninsula in air arriving after an overwater trajectory of about 1 km : observations are presented in Table 1 and Fig. 8. On 8 February 1963 steaming persisted from patches of water until the air temperature was only 0·4°C lower than the threshold value (Fig. 8). On 21 December 1963 steaming persisted until this difference was 0·9°C.

TABLE 1. Observation of the transition from steaming to no steaming

(a) 8 Feb. 1963. Surface temperature − 1·7°C ± 0·1. Salinity 31·5‰. Wind 7-10 m sec⁻¹ NNW. Temperature observations made at a height of 10 m

Time (LST)	Relative humidity (%)	Air temperature necessary for steaming (°C)	Measured air temperature (°C)	Difference (°C)	Steaming characteristics
0705	48·5	− 14·9	− 14·0	+ 0·9	none
0715	49·5	− 14·8	− 14·2	+ 0·6	none
0730	48	− 14·9	− 14·9	0	none
0745	45	− 15·1	− 15·0	+ 0·1	none
0810	44·5	− 15·2	− 15·1	+ 0·1	none
0815	41	− 15·4	− 15·6	− 0·2	none
0850	43·5	− 15·3	− 16·1	− 0·8	faint, widespread
0930	45·5	− 15·1	− 16·3	− 1·2	widespread, 70 cm
0950	44·5	− 15·2	− 16·4	− 1·2	1 m
1010	46·5	− 15·1	− 16·6	− 1·5	1 m
1030	43·5	− 15·3	− 16·1	− 0·8	fainter now
1050	49	− 14·9	− 15·9	− 1·0	faint, widespread
1100	45	− 15·1	− 15·7	− 0·6	in patches
1105	39·5	− 15·5	− 15·9	− 0·4	in patches
1115	43	− 15·3	− 15·6	− 0·3	in patches
1125	47·5	− 15·0	− 15·5	− 0·5	in patches
1135	48·5	− 14·9	− 15·3	− 0·4	very faint patches
1145	46	− 15·1	− 15·4	− 0·3	none
1200	46	− 15·1	− 15·1	0	none
1210	52	− 14·7	− 14·5	+ 0·2	none

(b) 21 Dec. 1963. Surface temperature 0·0 ± 0·3°C. Salinity 31·5‰. Wind 5-7 m sec⁻¹ NW. Temperature observations made at a height of 2 m

Time (LST)	Relative humidity (%)	Air temperature necessary for steaming (°C)	Measured air temperature (°C)	Difference (°C)	Steaming characteristics
0945	83	− 9·6	− 12·2	− 2·6	widespread, 1 m
0955	73	− 11·0	− 12·3	− 1·3	fainter
1000	83	− 9·6	− 12·6	− 3·0	1 m
1020	85	− 9·3	− 12·8	− 3·5	
1035	83	− 9·8	− 12·0	− 2·2	50 cm
1040	83	− 9·8	− 11·7	− 1·9	50 cm
1045	77	− 10·5	− 11·7	− 1·2	faint patches
1050	81	− 9·9	− 11·0	− 1·1	none
1052	81	− 9·9	− 11·1	− 1·2	faint patches
1055	82	− 9·8	− 11·3	− 1·5	patches
1056	83	− 9·6	− 11·1	− 1·5	patches
1057	81	− 9·9	− 11·5	− 1·6	faint patches
1059	81	− 9·9	− 11·9	− 2·0	patches
1100	74	− 10·9	− 11·8	− 0·9	none
1103	80	− 10·0	− 11·2	− 1·2	none
1114	81	− 9·9	− 11·5	− 1·6	faint patches
1117	74	− 10·9	− 11·7	− 0·8	none
1120	71	− 11·2	− 11·7	− 0·5	none
1125	66	− 11·8	− 11·4	+ 0·4	none
1130	74	− 10·9	− 10·2	+ 0·7	none

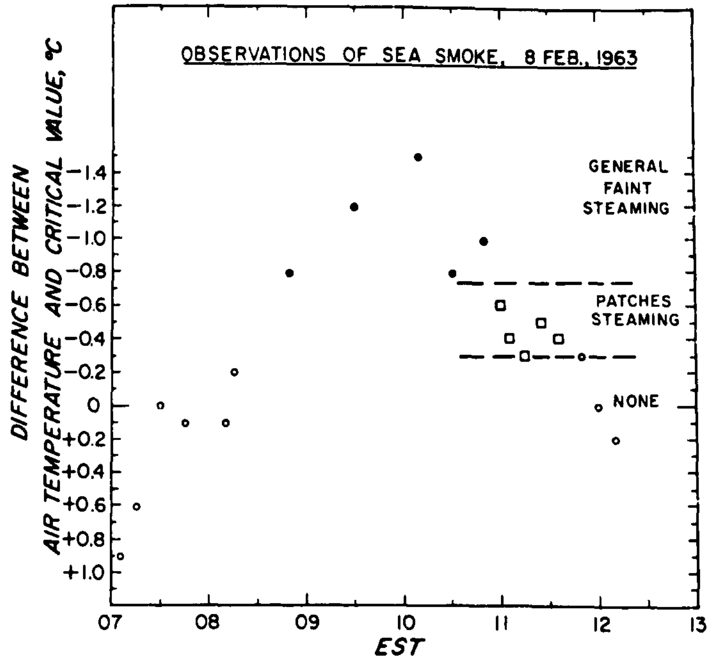

Figure 8. Observations of faint steaming, 8 February 1963 (author).

The steaming patches of sea water were presumably slightly warmer than their surroundings being produced by the upwelling induced by tidal currents and the discharge of effluents : no measurements were made of these surface temperature variations, but because of the highly stirred condition of the sea at these times it seems unlikely that they were larger than one- to three-tenths of a °C.

The data presented in the previous paragraphs indicate that, as employed to determine the conditions in which steaming occurs, the theory is an excellent first approximation. An improvement on it will need to recognize that for steaming to be apparent the conditions must have developed *beyond* the threshold stage. Thus (i) a certain minimum liquid water content (order 0·01 g m^{-3}) must be condensed out in order to provide sufficient visual contrast, and (ii) this condensed water must be raised (in turbulent fluctuations) to a certain minimum height above the water surface (order 10 cm). Condition (i) implies that for saline/fresh water the air must be approximately 0·4°C/0·7°C colder than the just no steaming value. Condition (ii) is dependent on the wind and thermal structure in the air in a way which has yet to be investigated.

Acknowledgments

This work was made possible by the support of the Office of Naval Research under grant No. 2196 (00) and is Contribution No. 1406 from the Woods Hole Oceanographic Institution. The writer wishes to record his appreciation for the help of N. P. Fofonoff who translated the valuable paper of G. M. Bashkirova and P. N. Krasikov and who also indicated the scope of other Russian works; also to A. H. Woodcock for many illuminating discussions and access to unpublished data.

Appendix

A list of terms for the clouds of ' steam ' rising from warm water into cold air : vapour (laymen); black frost, white frost (N. Atlantic fishermen); sea mist, sea smoke, Arctic sea smoke, Arctic smoke; frost smoke and barber (crystalline.) Steam mist, autumn mist, water smoke; cold air advection fog, evaporation fog and mixing fog.

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* Reports of presence or absence of sea smoke used in compiling Fig. 6.

Pub. No. 10, Great Lakes Research Division, The University of Michigan, 1963

THE BIOLOGICAL INTEGRATORS

SOME BIOLOGICAL CHARACTERISTICS OF THE MARINE ENVIRONMENT*

Bostwick H. Ketchum
Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

For several reasons I have elected to talk about the field of phytoplankton biology, to the neglect of many other fascinating aspects of marine biology. First is the selfish reason that my own interests have been in this area for years. More importantly, the problems of plankton biology do not recognize the salt barrier—they are identical in the lakes and in the sea. Though the species may change the problems remain the same. Perhaps the most important reason for this selection, however, is that an understanding of plankton population growth and distribution must be firmly based upon all disciplines in oceanography and limnology. The currents of the oceans carry these populations of "drifters" about and modify and determine the centers of abundance. Turbulent mixing in the oceans and the depth of the thermocline determine whether the phytoplankton are in the illuminated zone of the sea enough of the time to permit them to produce an excess of living matter over their immediate metabolic needs. Turbulent mixing also may bring elements essential to plant growth to the thin surface layers where they may be used. The physics of light penetration through the waters determines the depth to which plants may grow and the chemistry of the water establishes its fertility. Geologically the character of many marine sediments is determined by the plankton history of the overlying waters and conversely the little understood exchanges of elements between the sediments and the overlying water must in the long run influence, if not determine, the distribution and proportions of elements in the sea. This list could be expanded but should suffice to suggest that a plankton ecologist could profit by divine omniscience in his approach to his problems.

The "where, how much and why" of plankton populations is determined by both biotic and abiotic factors in the environment. Today I shall emphasize the latter, or abiotic, factors. Within the last few decades we have developed some understanding of the quantitative influence of these factors on plankton populations but we still do not know enough to predict or explain which species may be present in abundance at any time or place.

The abiotic environmental factors can be grouped into three categories, two of which essentially determine the quantitative character—the biomass or how much of a given population can be produced—and the third is primarily important in determining the quality—the whether and where of populations.

The first, and certainly a fundamental factor, is energy. For autotrophic plants the source of energy is, of course, light; for chemoautotrophic organisms energy-rich inorganic chemicals, ammonia for example, suffice; for heterotrophs and saprophytes chemical energy in the form of organically fixed carbon is essential. Ultimately, however, the energy flow of the entire community is derived from sunlight. The fundamental characteristic of the energy

*Contribution No. 1432 from the Woods Hole Oceanographic Institution. Some of the studies discussed herein have received financial support from the U. S. Atomic Energy Commission (AT(30-1)-1918), from the Office of Naval Research (contract Nonr 2196(00)) and from grants by the National Science Foundation.

system is that it is a "one way street." Energy enters the system, is accumulated in organic compounds and passed along among successive members of the food chain, being dissipated all along the way. The ecosystem is kept in operation by continuous additions of energy which, once dissipated, cannot be returned to the system.

The second fundamental factor, nutrients, determines the fertility of the water. Nutrients differ from energy since they may be used and reused in the development and succession of populations. Thus we may talk of the cycles of phosphorus, of nitrogen or of carbon in the sea in contrast to the one-way street for energy. This recycling of elements introduces many difficult problems in evaluating plankton production and the fertility of waters which are still awaiting solution. For example, in many oceanic waters vertical turbulence cannot bring to the surface waters sufficient nutrients to maintain the observed rates of photosynthesis. *In situ* recycling of nutrients is essential to the system. We have not yet devised ways to measure this directly, we can only say it must be so or the system could not work.

The third abiotic character of the environment can be termed the conditioning factors. These are not used up in the growth or development of populations, but they are important in determining whether a given species can exist at a given time and place. In the sea these include the salinity and temperature of the water and, where present, pollutants, either natural or derived from man's activities. Nature is remarkable in that species have evolved which can exist in all parts of the ocean, and in general the biomass produced is more a function of the environment than of the species. Consequently these conditioning factors generally do not determine quantity, but they have a great influence on quality—on which species can survive at a given location. These factors thus determine geographical distribution so that we may speak, for example, of boreal or tropical populations and may differentiate between estuarine, neritic and pelagic species. Conditioning factors will not be discussed in detail in this paper since even a brief summary would exhaust both the space and time available.

In summary, then, the energy flow and the supply and recycling of nutrients are the principal determinants of the biomass which may be produced in any oceanic environment; the species composition of the population is certainly dependent upon energy and nutrient supply but may also be modified by other characteristics of the environment.

Solar radiation, the source of energy for the ecological cycle in the sea, varies daily, seasonally and geographically. The total energy impinging upon the sea surface and available for photosynthesis is a function both of the intensity and the length of daylight, as modified by the average cloudiness of the particular area being considered. The greatest average annual intensity is in the equatorial regions where the sun is always near the zenith and the daylength never differs much from twelve hours. The greatest monthly average energy is found in temperate latitudes in the summer, when the total energy available during the day may be nearly double the amount available at the equator. For example, in June the maximum daily radiation at about 40° North latitude is .360 g-cal/cm^2/min and in January at 30° South the average is .452 g-cal/cm^2/min (Sverdrup *et al.* 1942, p. 103). Even at 60° North where the sun never shines in December, the total energy reaching the surface in June, when the sun never sets, is about equal to the energy reaching the equatorial waters because the lower intensity continues throughout 24 hours of the day.

After entering the sea, the intensity of light is attenuated by absorption in the water by colored materials in solution and by organisms. It is also attenuated by the scattering from particles both living and dead. As a result

of these various processes, light at different wave lengths decays at different rates and since the attenuation is logarithmic only a thin surface film of the oceans is suitable for photosynthesis (Clarke 1939). As a general rule of thumb, ecologists have accepted 1% of the incident radiation as being the intensity at which the photosynthesis of the plant population exactly balances its own needs for respiration. Where this occurs in the water column is called the compensation depth. In the clearest ocean waters this may be found at about 100 m, but in more turbid coastal or harbor waters, it may be considerably less, perhaps one or a few meters.

To the physiologist it would seem more logical to specify an energy unit rather than a percentage of incident light to define the compensation depth in the ocean. This has indeed been done (for example, Jenkin 1937). In the natural environment, however, evidence is accumulating that the phytoplankton may adapt to different light conditions (Steemann Nielsen and Hansen 1959a; Ryther and Menzel 1959). Under relatively stable conditions it appears that the phytoplankton exposed to a low light intensity can use this intensity more effectively than those exposed to higher light intensities. Presumably this is accomplished by the accumulation of additional concentrations of chlorophyll *a* within the cell.

The clearest ocean waters are clear partly because the phytoplankton population there is very sparse. Where the phytoplankton population is dense, the cells are very effective in absorbing and scattering the incident light. Steemann Nielsen (1954) has observed that there is an inverse relationship between the depth of the compensation layer and the total amount of photosynthesis that can take place within the water column. Clearly only the light which is absorbed by chlorophyll or the accessory pigments of the algal cell is effective in photosynthesis, and where practically all of the light is absorbed by chlorophyll total photosynthesis will be at a maximum. There are, of course, sources of turbidity other than the living phytoplankton cells in the water so that the relationship between turbidity and total photosynthesis is not a simple one. Riley (1956a) has developed equations for calculating, under various environmental conditions, the fraction of the incident radiation which is absorbed by the water, by plant pigments, and by particles.

The vertical distribution of photosynthesis in the sea is further complicated by the fact that high light intensities may be inhibitory. An example of the variation of photosynthesis at various depths (light intensities) throughout the day is shown in Figure 1 (Ryther 1956). These are theoretical curves, derived from measurements of the photosynthesis of phytoplankton algae as a function of incident light intensity. They assume a phytoplankton population uniformly distributed throughout the euphotic zone, a distribution rarely if ever observed in nature. On the brightest day, when the total incident surface radiation was about 800 g-cal/cm^2, the mid-day inhibition of photosynthesis by intensities greater than 25% of the incident radiation is clearly demonstrated. During the darkest day of the year, when the total surface radiation was about 25 g-cal/cm^2/day, the surface radiation never reached inhibitory intensities and the maximum rate of photosynthesis was only about 50% of the rate which could have been achieved by this hypothetical population at saturation light intensity.

This brief and incomplete discussion of the factors involved in the relationship between incident light intensity and phytoplankton productivity should suffice to indicate the type of variables which must be considered in evaluating any given situation. These variables certainly include the incident solar radiation, the attenuation by absorption or scattering in the sea, the physiological state of the organism (see below), and the vertical distribution of the phytoplankton in the water column.

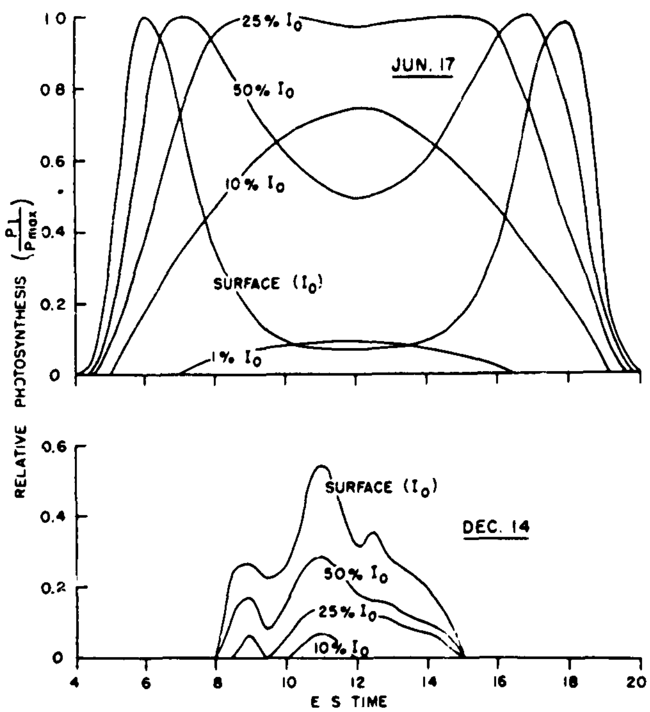

Fig. 1. Calculated relative photosynthesis at various fractions of incident radiation on the brightest (June 17) and darkest (Dec. 14) day in 1954. (After Ryther 1956)

The importance of nutrients in determining the rate of photosynthesis and the productivity of phytoplankton populations has long been recognized and was recently reviewed by Ketchum *et al.* (1958). Using cultures of a marine flagellate, Ryther (1954) demonstrated that the rate of photosynthesis declined rapidly as the cells exhausted the supply of nutrients in the medium. When cell division stopped because of exhaustion of nutrients, photosynthesis and respiration proceeded at approximately equal rates. The results of a similar experiment are shown in Figure 2 (Ketchum *et al.* 1958). A pure culture of *Dunaliella euchlora* was grown in a medium deficient in phosphorus, and periodically the phosphorus and chlorophyll *a* content of the cells was determined and the net and gross photosynthesis was measured by the light and dark bottle oxygen method. The cells rich in phosphorus gave a net:gross photosynthesis ratio of about 0.9, characteristic of healthy natural populations (see Steemann Nielsen and Hansen 1959b). As the ratio of phosphorus to chlorophyll *a* fell below about 25 μg atom/mg the net:gross photosynthesis ratio decreased drastically. These examples demonstrate that nutrient deficiencies can develop in cultures and seriously modify the photosynthesis of the algae.

The sea is, however, a dynamic system, and conclusions concerning productivity of a given area cannot be reached by the measurement of the concentration of available nutrients in the environment at any given time. In the past many efforts have been made to correlate the concentration of nutrients and the biomass of the phytoplankton population, but unless this is done on a continuing time basis the efforts are doomed to failure. In a confined volume of water in the laboratory one is justified in the assumption that,

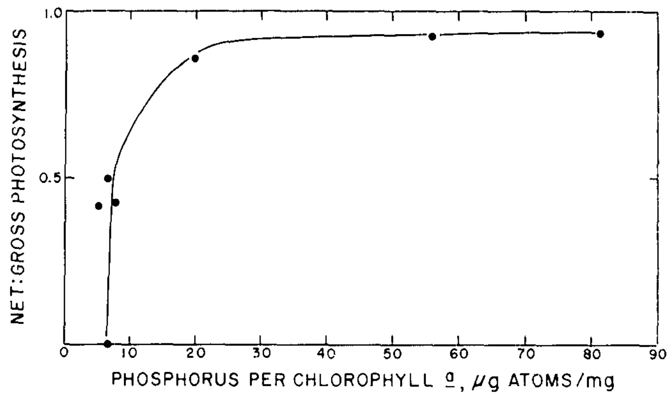

Fig. 2. The ratio of net:gross photosynthesis as a function of the ratio of phosphorus:chlorophyll *a* in the cells of a culture of *Dunalicella euchlora*.

within limits, the quantity of essential nutrients added to the water at the start of an experiment will determine the total size of the population which can be developed. In the sea, on the other hand, organisms are constantly dying and settling out of the euphotic zone carrying essential nutrient elements with them, turbulent mixing is constantly adding to the euphotic zone nutrients from the richer deep waters, bacterial decomposition of dissolved organic material is constantly taking place and liberating nutrients, the phytoplankton are being eaten by zooplankton which may excrete directly the essential elements in a form suitable for immediate phytoplankton reutilization but may also die and sink, carrying nutrients out of the surface layers. In some areas of the world, the wind and current systems may produce upwelling of water from intermediate depths, thereby enriching the surface layers. In different parts of the world these processes will vary in their importance in controlling and maintaining the continuation of photosynthesis and the total productivity of the water column.

The distribution of areas of high productivity in the world is shown in Figure 3 (Emery 1963). From the previous discussion of the distribution of light intensity it is clear that light alone is inadequate to explain any such distribution. These areas of high productivity are largely related to upwelling areas where water from intermediate depths is brought to the surface and made available to the phytoplankton. On the west coasts of continents the persistent offshore winds, particularly in the belt of prevailing trade winds, move the surface water away from the coast and allow water from intermediate depths (200 to 300 m) to reach the surface. In the equatorial regions diverging current systems similarly enrich the near surface layers. Sette (1955) has described the sequence of upwelled nutrients and the develonment of phytoplankton and later of the animal members of the food chain in the equatorial region of the Pacific Ocean. The near polar regions are also areas of upwelling which are related to the current systems of the area.

It is significant that the large central basins of the oceans are not highly productive areas. The most extensive surveys of productivity of a mid ocean area have been carried out in the Sargasso Sea near Bermuda (Menzel and Ryther 1960, 1961; Ryther and Menzel 1960, 1961). In such areas there is a marked stability of the water column throughout the year so that the transport of nutrients from the deeper water to the surface is small, and they do not have the winter deep mixing which is typically found in the temperate latitudes. In the Sargasso Sea, Riley (1957) and Menzel and Ryther (1960) have concluded

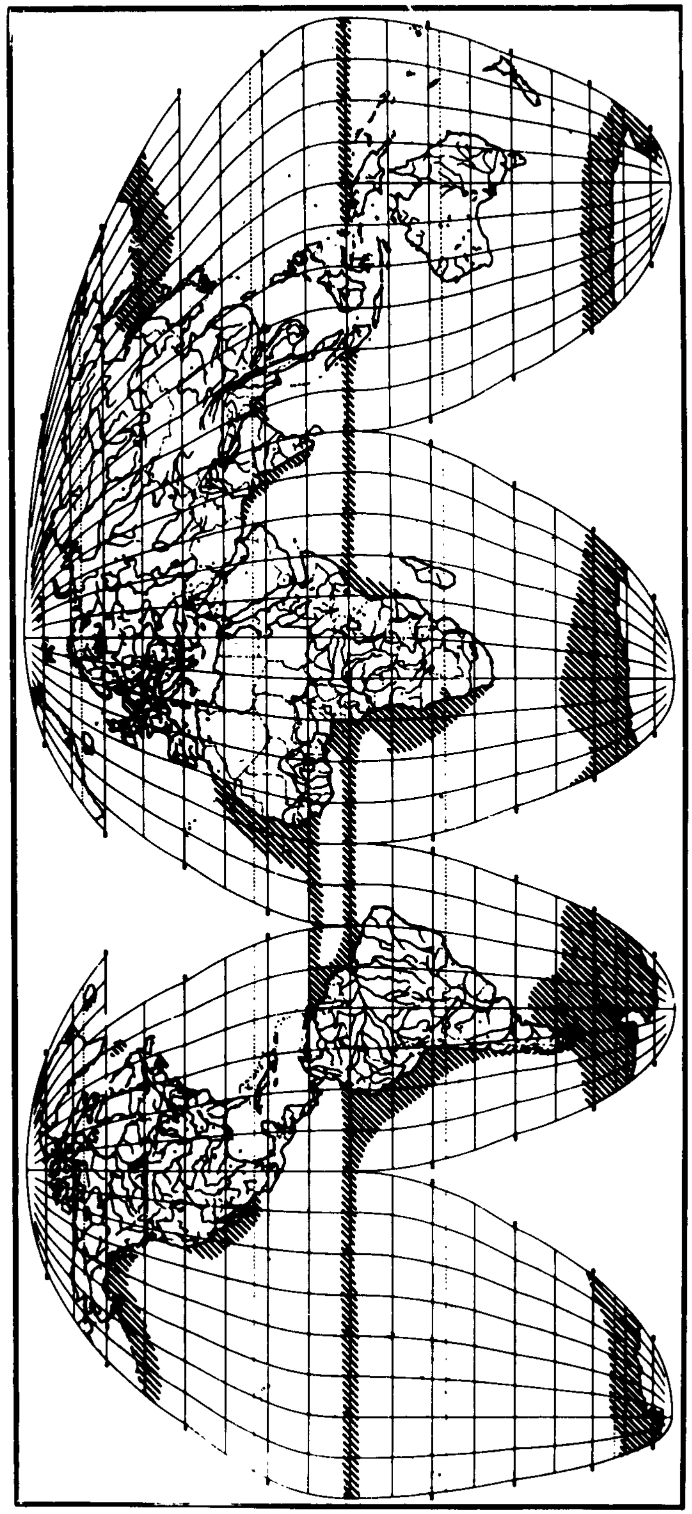

Fig. 3. Oceanic areas of high phytoplankton productivity. (After Emery 1963).

Table 1. The Annual Productivity, Phosphorus Requirement and Utilization and Recycling of Phosphorus in the Coast Waters of the New York Bight (September 1956 - September 1957, 371 days)

Location of stations	No of stations	Depths (m)	Productivity (mg $C/m^2/yr$)	Phosphorus requirement (mg A $P/m^2/yr$)	Observed P utilization (mg A $P/m^2/yr$)	Recycling ratio
Montauk Pt. to Hudson Canyon	6	37-45	2190	182.5	38.5	4.74
Off Barnegat, N.J.	5	25-40	2210	184.0	34.9	5.33
Off Montauk Pt.	6	63-75	1795	149.5	42.9	3.48

that the major biological activity is confined to the first 300 m or so of the water column. The major part of the productivity must be maintained by the recycling of nutrients *in situ.*

Some estimates have been made of the relative importance of the total available supply of nutrients, of *in situ* recycling, and of vertical mixing in maintaining the productivity of certain areas. For example, the data of Redfield *et al.* (1937) indicate that only about 2% of the total phosphorus cycle between May and November was reflected by changes in the quantity of inorganic phosphorus in the euphotic zone whereas about 73% was supplied by vertical mixing and about 25% by regeneration (Ketchum 1947). Steele (1956) found about 18% of the summer production was attributable to changes in phosphate in the euphotic zone, and 82% was contributed by vertical mixing. In Long Island Sound, Riley's (1956b) estimate of gross production would require about 10 times the total amount of phosphorus in the water at its maximum concentration (Riley and Conover 1956).

We have made a more recent estimate in the coastal waters south of Long Island and some results are summarized in Table 1. These waters are thoroughly mixed in the winter time when practically all of the phosphorus is present as available dissolved inorganic phosphate. Throughout the year estimates of productivity were made and have been described by Ryther and Yentsch (1958). At these same times measurements of the phosphorus content of the water were also made. At various stations the amount of organic carbon fixed was summed throughout the year and the requirement for phosphorus was computed from this assuming that one atom of phosphorus was needed for every hundred atoms of carbon fixed (see Redfield 1934). The observed utilization of phosphorus was also evaluated according to the method of Redfield *et al.* (1937) which gives a minimum estimate but does take account of vertical transport of the element and of some recycling *in situ.* The requirement for phosphorus greatly exceeds the available supply at any one time and the recycling ratio indicates that each atom of phosphorus must pass completely through the biological system four or five times a year in the euphotic zone in order to produce the amount of carbon fixation observed. In the Sargasso Sea, Menzel and Ryther (1960) found that the observed primary productivity should completely utilize the nutrients in the upper 400 meters of water but that it proceeded with no substantial changes in the observable concentrations. They concluded that in this area, except for a brief period in the spring, the entire productivity cycle must be supported by *in situ* recycling of nutrients which must proceed comparatively rapidly in the warm waters off Bermuda.

Over the last few decades we have learned a great deal about the environmental control of the productivity of marine waters so that one can speak with some confidence of the causes for highly productive or impoverished areas. Much, however, remains to be done. There is still controversy about the methods for measurement of photosynthesis of natural populations and the definition of exactly what characteristic of the process each method is measuring is still unresolved. We cannot yet determine directly the rate of recycling of nutrients which must be so important in maintaining productivity of some areas. We cannot determine the respiration of phytoplankton in mixed plant and animal communities, which we must be able to do in order to answer some of our questions concerning productivity and the food available for the next trophic level. We do not know how useful various plant species are as food. Which are weeds and which are the valuable pastures of the sea? In every locality which has been adequately studied there is a seasonal sequence with one species being dominant at one time to disappear and be replaced by another. This sequence we can describe but cannot yet explain or predict. These and many other questions keep plankton biology an interesting and challenging field of study.

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